$^{40}Ar/^{39}Ar$ laserprobe dating of mylonitic fabrics in a polyorogenic terrane of NW Iberia

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Abstract: The tectonothermal evolution of a polyorogenic terrane in the Variscan belt of NW Spain has been constrained by $^{40}Ar/^{39}Ar$ laserprobe incremental heating experiments on mylonitic fabrics developed in major structures. Transitional levels between HP–HT and IP upper units in the Órdenes Complex where metamorphic and structural records demonstrate two cycles of burial and exhumation were selected for dating. Two groups of ages have been defined: (1) Silurian–Early Devonian, obtained from mylonites of the Fornás extensional detachment, here considered as the minimum age for the start of tectonic exhumation of the HP–HT units and an upper age-limit for the HP–HT event itself; (2) Early to Mid-Devonian, from structures related to the Variscan convergence in the area, which include top-to-the-east thrusts and extensional detachments. A single, younger Carboniferous age obtained from the uppermost allochthonous sequences possibly reflects the final stages of emplacement of the allochthonous complexes. Our data indicate a polyorogenic character for a part of the Iberian allochthonous complexes, including Variscan (*sensu stricto*) and Early Variscan convergence, as well as an older, Early Palaeozoic cycle.

It is commonly accepted that terrane accretion is a first-order mechanism for plate growth. Deciphering the sequence of amalgamation and the origin of every terrane involved in an orogenic belt is a complex task and requires coordinated work on structural geology, petrology and geochronology. The European Variscan belt and its North American counterpart, the Appalachian belt, are good examples for the study of such tectonic processes (van Staal *et al.* 1997; Martínez Catalán *et al.* 2002; Winchester & The PACE TMR Network Team 2002).

The Variscan belt of NW Iberia is characterized by the existence of five allochthonous complexes preserved as megaklippen and resting upon autochthonous and parautochthonous sequences (Martínez Catalán *et al.* 2002). The Órdenes Complex, in Spain, is the largest of them, and consists of a stack of allochthonous units, including ophiolites defining a suture, and suspect terranes of peri-Gondwanan provenance (Fig. 1). At present, the origin and detailed evolution of these units is controversial. Their position during the Early and Mid-Palaeozoic and their interactions with other well-known terranes such as Avalonia and European correlatives (Winchester & The PACE TMR Network Team 2002) remain unclear.

Several tectonothermal and geochronological studies (Schäfer *et al.* 1993; Dallmeyer *et al.* 1997; Ordóñez Casado *et al.* 2001; Fernández-Suárez *et al.* 2002) in the upper units of the allochthonous complexes support the existence of a pre-Variscan record (Abati *et al.* 1999; Fernández-Suárez *et al.* 2002; Santos *et al.* 2002) followed by Early Variscan ages and a Variscan (*sensu stricto*) evolution similar to that of the NW Iberian autochthon (Dallmeyer *et al.* 1997). However, the transition between pre-Variscan and Variscan evolutions remains obscure, and this includes the age of high-pressure and high-temperature metamorphism. To clarify this transition, we present new $^{40}Ar/^{39}Ar$ laserprobe data from single- and multi-grain experiments on mylonite metabasites and schists from the upper units of the Órdenes Complex. The samples were collected after a detailed structural and metamorphic analysis of a key section of the upper units: the transition between high-pressure and high-temperature metamorphism (HP HT) and intermediate-pressure (IP) upper units (Fig. 1). To test the relationships between structural and tectonic events only mylonites from well-constrained structures were sampled, including extensional detachments and ductile thrusts. The results reveal intimate relationships between structural and tectonic processes and provide a more complete understanding of the polyorogenic tectonothermal evolution of this peri-Gondwanan terrane.

Geological setting

The Órdenes Complex consists of a stack of thrust sheets, overprinted by extensional detachments, upright folds and faults (Martínez Catalán *et al.* 2002). Three main tectonometamorphic units form the Órdenes Complex; these are, from bottom to top, the basal, ophiolitic and upper units (Fig. 1). The basal units are interpreted as belonging to the margin of Gondwana (Martínez Catalán *et al.* 1996). They contain HP low/intermediate *T* relicts related to the subduction of the margin during the Variscan orogeny.

The upper units can be subdivided into HP HT units below and intermediate-pressure (IP) units occupying the uppermost position. At the top of the IP units a package of low-grade graywackes can be differentiated (Betanzos unit; Fig. 1). Contacts between units are faults, often extensional detachments

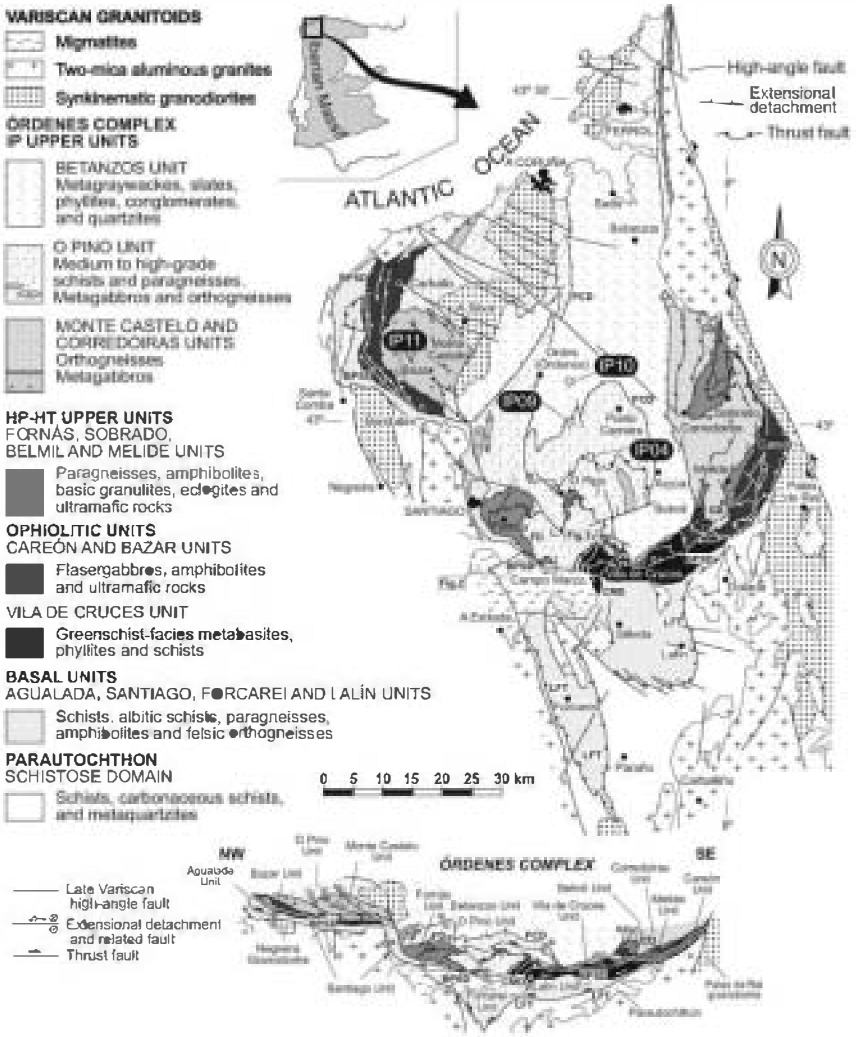

Fig. 1. Geological map and cross-section of the Órdenes Complex (NW Spain) showing the various units and the locations of samples IP-04 and IP-09 to IP-11 for $^{40}Ar/^{39}Ar$ geochronology. The other samples are located in the detailed maps of Figures 2 and 3. FD, Fornás detachment; CD, Corredoiras detachment; PCD, Ponte Carreira detachment; BPSD, Bembibre Pico Sacro detachment; CMD, Campo Marzo detachment; LFT, Lalín Forcarei thrust.

(Fig. 1), the most important of which are the Fornás, Corredoiras and Ponte Carreira detachments (Díaz García *et al.* 1999*b*; Castiñeiras 2003; Gómez Barreiro 2004). These structures attenuated the initial orogenic pile, putting together rocks that initially were situated in different positions, as can be derived from the tectonothermal and geochronological evolution (e.g. Castiñeiras 2003; Fernández-Suárez *et al.* 2003; Gómez Barreiro 2004). The Monte Castelo gabbro and the Corredoiras orthogneiss are good samples of arc magmatism in the IP units (Fig. 1). They are located at the bottom of the IP units and recorded a comparable tectonothermal evolution (Abatí 2002; Castiñeiras 2003; González Cuadra 2005). The metasedimentary and igneous protoliths of the HP–HT units are equivalent to those of the IP set (Arenas & Martínez Catalán 2002; Fernández-Suárez *et al.* 2002, 2004; Gómez Barreiro 2004). Higher P–T conditions and the presence of ultramafic rocks indicate that the HP–HT units occupied a lower position in the pile. Any indication of a possible suture between IP and HP–HT sets is lacking, so that they are considered as parts of the same tectonostratigraphic terrane in spite of their different metamorphic evolution. In fact, the two sets show a complementary metamorphic evolution, with two cycles of burial and exhumation, indicating a thermobaric convergence when approaching the transition zone, that is, the detachment separating them (Fornás detachment; Figs 1, 2 and 4). These zones of contact may show a subsequent record of overprinting events and reworking (Gómez Barreiro 2004). The present study aims to decipher the tectonothermal record of these transitional levels, a crucial point in understanding the stacking of the allochthonous units through time, and hence the orogenic cycles involved.

From a geodynamic perspective the upper units are considered exotic terranes of magmatic arc affinities (Andonaegui *et al.* 2002; Abati *et al.* 2003), which are separated from the basal units by an orogenic suture (Martínez Catalán *et al.* 1996, 1997, 1999). The ophiolitic units define that boundary and include different sets of rocks with a suprasubduction character and a protolith age of 395 ± 2 Ma (Díaz García *et al.* 1999*a*; Pin *et al.* 2002; Sánchez Martínez *et al.* 2005).

Understanding the geodynamic context and evolution of the allochthonous terranes requires a detailed analysis of metamorphic and structural relationships of the HP–HT and IP sets, constraining the age of the main geological features. This work has been focused on the SW part of the Órdenes Complex, where good exposures of the boundary between HP–HT and IP units exist (Gómez Barreiro 2004).

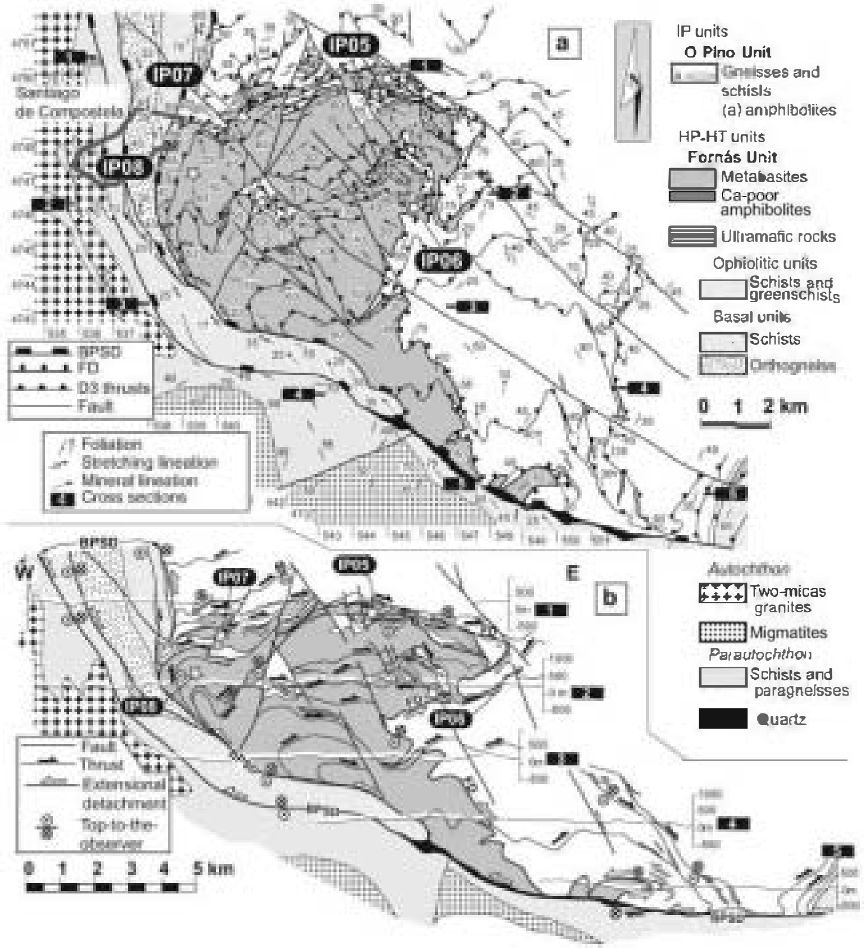

Fig. 2. Geological map and composite cross-section of the Fornás unit. Locations of samples IP-05 to IP-08 for $^{40}Ar/^{39}Ar$ analysis are shown.

Previous geochronological data

Previous dating experiments in the upper units have tried to establish the age of the HP–HT metamorphic event, the magmatism, and sediment age and provenance (see Dallmeyer *et al.* 1997; Abati *et al.* 1999; Ordóñez Casado *et al.* 2001; Fernández-Suárez *et al.* 2003, and references therein). To summarize results, data from the upper units have been grouped into four age intervals, with reference to related data in other units (see Fig. 6), as follows.

Pre-520 Ma

This group encompasses two types of ages. The first is related to the Early and Late Proterozoic and Early Palaeozoic provenance of detrital sediments in the upper units (Peucat *et al.* 1990; Ordóñez Casado *et al.* 2001; Fernández-Suárez *et al.* 2003). The second derives from conventional U–Pb zircon dating in meta-igneous rocks whose protolith ages cluster around 500 Ma. Upper intercept ages of 1.95–2.5 Ga have been found in orthogneisses of the Órdenes and Morais Complexes (Kuijper 1979; Dallmeyer & Tucker 1993), whereas granulite- and amphibolite-facies rocks sampled in units derived from mafic protoliths have yielded upper intercept ages of 2.6–2.8 Ga (Peucat *et al.* 1990; Santos Zalduegui *et al.* 1996).

520–480 Ma

Widespread Cambro-Ordovician magmatism at 520–500 Ma has been documented in the upper units of the allochthonous complexes (Kuijper 1979; Van Calsteren *et al.* 1979; Peucat *et al.* 1990; Dallmeyer & Tucker 1993; Schäfer *et al.* 1993; Abati *et al.* 1999; Ordóñez Casado *et al.* 2001; Fernández-Suárez *et al.* 2002, 2004; Santos *et al.* 2002). In the Órdenes Complex, both arc-type magmatism and a related thermal event have been recognized and dated at *c.* 500–495 Ma (Abati *et al.* 1999, 2003). Furthermore, detrital zircon ages in the uppermost, low-grade units (Betanzos unit, Fig. 1) permit us to establish a maximum limit for the depositional age of the sedimentary protolith of *c.* 480 Ma (Fernández-Suárez *et al.* 2003).

430–390 Ma

The meaning of geochronological data in this interval representing the pre-Variscan–Early Variscan transition is unclear.

In the HP–HT units, ages around 395 Ma have been interpreted as dating the HP–HT event (M_1) (Dallmeyer *et al.* 1991; Ordóñez Casado *et al.* 2001). As a consequence, a polyorogenic character for the upper units has been ruled out by some workers (Ordóñez Casado *et al.* 2001). However, a few Silurian ages *c.* 430–425 Ma have been obtained by $^{40}Ar/^{39}Ar$ and U–Pb conventional methods (Dallmeyer *et al.* 1997; Fernández-Suárez *et al.* 2002), and recent studies involving U–Pb ion microprobe analyses on zircons have tried to establish the different population ages formed in the same rocks along the entire P–T path (Fernández-Suárez *et al.* 2004). Gabbros, mafic granulites and migmatitic leucosomes were investigated (Fernández-Suárez *et al.* 2004), and it was shown that new zircon was growing since at least 410 Ma, in close relation to partial melting. Also, it is widely accepted that all HP–HT units underwent partial melting after having reached their pressure peak (Arenas 1991; Arenas & Martínez Catalán 2002). Consequently, this 430–390 Ma time interval probably comprises a mixture of tectonic events, including the exhumation of the HP–HT rocks after the pressure peak.

Post-390 Ma

This group represents the Variscan and Early Variscan evolution of the upper units and correlates well with the evolution of the rest of the allochthonous units and the autochthonous sequences (Peucat *et al.* 1990; Dallmeyer *et al.* 1991, 1997; Abati *et al.* 1999). The first tectonic fabric developed in ophiolitic units has been dated at *c.* 385 Ma ($^{40}Ar/^{39}Ar$ in hornblende concentrates; Dallmeyer *et al.* 1997), indicating that plate convergence was active during the Mid-Devonian.

Tectonothermal evolution of the Fornás, Arinteiro, and O Pino units

In the SW part of the Órdenes Complex, HP–HT units are represented by the Fornás and Arinteiro units (Figs 2 and 3), consisting mainly of amphibolites with minor peridotites, garnet metapyroxenites and paragneisses (Gómez Barreiro 2004). These units represent the same HP–HT sheet, tectonically dismembered during the emplacement of the allochthonous sequences onto the Variscan autochthon. The IP O Pino unit consists of pelitic and semipelitic schists and gneisses, with metagabbroic rocks and granitic leucosomes (Castiñeiras 2003). The Fornás, Arinteiro, and O Pino units have recorded seven correlative tectonothermal stages (Fig. 4), as follows.

D_0–M_0

This is the oldest thermal record, preserved only in some pelitic gneisses of the O Pino unit, where mesoscopic veins with And–Qtz (mineral abbreviations after Kretz 1983) appear. These veins have been related to low-pressure heating in a volcanic-arc setting where magmatism was widespread (Castiñeiras 2003). Evidence of this event in the Fornás and Arinteiro units is preserved as a few meta-igneous relics.

D_1–M_1

Both units reached metamorphic peak conditions during this event. In the O Pino unit, the equilibrium association at the metamorphic peak was Ky + Grt + Bt + Pl ± St + Qtz (9.7 ± 1 kbar and 650 ± 20 °C (Castiñeiras 2003)). Petrographic evidence suggests that the rocks were pressurized following an isothermal path and finally retrogressed, with M_0 and M_1 stages defining an anticlockwise P–T path (Fig. 4), as in other IP units

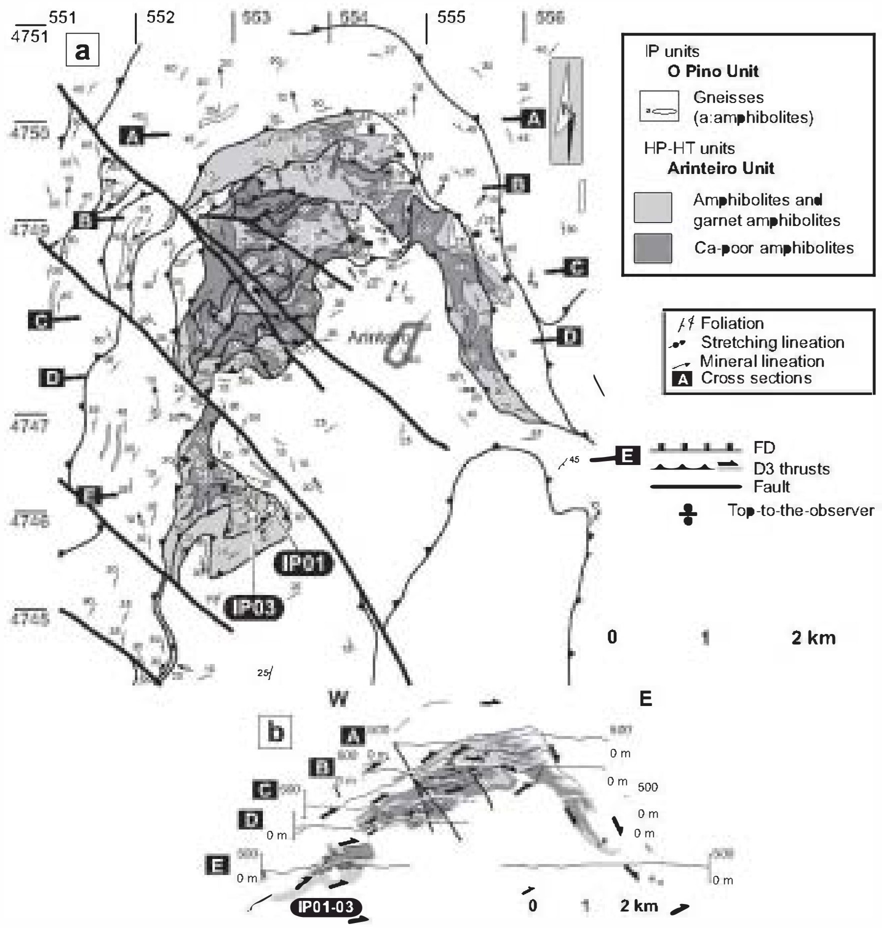

Fig. 3. Geological map and composite cross-section of the Arinteiro unit (HP–HT). Locations of samples IP-01 and IP-03 for $^{40}Ar/^{39}Ar$ analysis are shown.

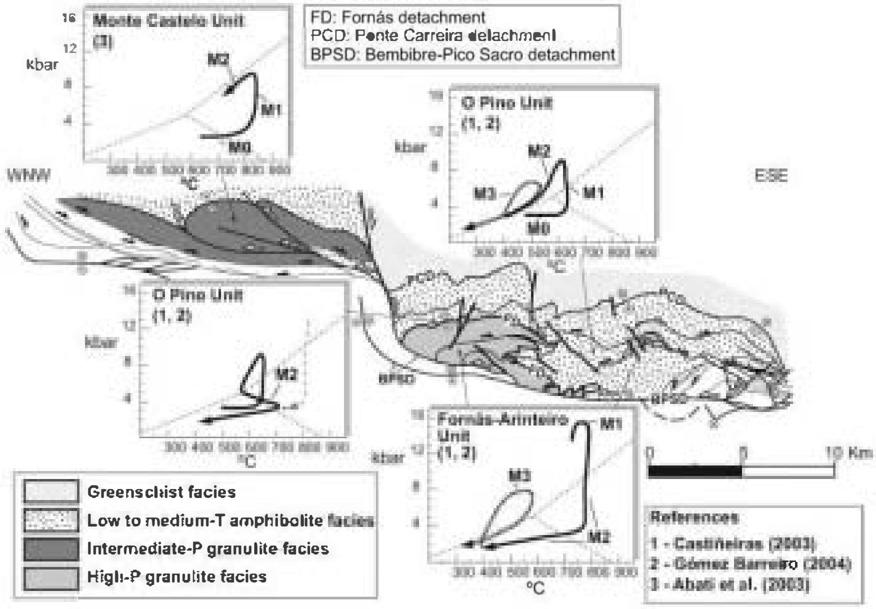

Fig. 4. Distribution of metamorphic facies in the upper units of the west of the Órdenes Complex, as seen in cross-section, and P–T paths for several units. (See text for a discussion of metamorphic stages (M_0, M_1, ...).) Dashed lines in P–T diagrams represent And–Ky–Sil stability fields and triple point after Powell & Holland (1990). Fault types as in Figure 1.

of the Órdenes Complex (Abati *et al.* 2003; Castiñeiras 2003). The regional foliation in the O Pino unit was formed at this stage.

In the Fornás and Arinteiro units, high-pressure granulite-facies assemblages were developed, with peak paragenesis formed by Cpx + Pl + Grt + Rt + Qtz in the mafic rocks, and marked by the growth of garnet after spinel in ultramafic rocks. Other HP–HT units of the Órdenes Complex reflect an equivalent P–T evolution (12–23 kbar and 750–850 °C (Gil Ibarguchi *et al.* 1990; Arenas & Martínez Catalán 2002). The HP–HT granulitic fabric was preserved only as decimetre-scale lenses surrounded by the decompressive D_2–M_2 fabric (Gómez Barreiro 2004).

D$_2$–*M*$_2$

The thickened pile led to gravitational instabilities, resulting in several major extensional detachments in the upper units (Gómez Barreiro 2004). The Fornás detachment is the contact between the HP–HT and the IP units (see Figs 2 and 3). A strong regional fabric was formed in the footwall to the detachment (Fornás and Arinteiro units), mainly developed under amphibolite-facies conditions. The P–T evolution shows isothermal decompression at over 700 °C where high-temperature assemblages were equilibrated (Cpx + Pl + Hbl) and partial melting occurred (Arenas 1991; Gómez Barreiro 2004). Later, isobaric cooling at low pressures (<4 kbar) led to a progressive substitution of Ca-rich amphibole by cummingtonite in the mafic rocks (Gómez Barreiro 2004). Tectonites developed during decompression were banded mylonites. With increasing strain as T decreased, deformation concentrated in the upper levels of the Fornás and Arinteiro units, where mylonites and ultramylonites developed in discrete shear zones. At greater strain, no evidence of compositional banding persists and a mixture of phases (Hbl–Pl and Cum–Pl) replaces it. All tectonites show a strong shape and crystallographic preferred orientation of amphibole, pyroxene, plagioclase and quartz, which defines a persistent NNW–SSE mineral lineation (Fig. 2). The kinematic analysis of the Fornás detachment reveals a general top-to-the-NNW sense of shear (Gómez Barreiro 2004). Taking into account microstructural features and petrofabric analyses, the conditions under which the deformation progressed could be bracketed between 750 °C and 300 °C (Gómez Barreiro 2004).

In the hanging wall to the Fornás detachment (O Pino unit), tectonothermal effects were restricted to the first few metres above the contact. The rocks were equilibrated under medium- to low-pressure and high-temperature conditions (Fig. 4). The main assemblages in pelitic gneisses were Grt + Bt + Kfs + Sil + Qtz + melt and Crd + Bt + Kfs + Qtz ± Grt + melt. Pressures <4 kbar and temperatures >650 °C are inferred from petrographic analysis based on the petrogenetic grid of Powell & Holland (1990) (Gómez Barreiro 2004). Synkinematic leucosomes intruded into the bottom of the O Pino unit, showing pegmatitic textures and granite–granodiorite compositions. Mylonitic bands were developed during D_2, with kinematic criteria showing a consistent top-to-the-NNW sense of shear and a NNW–SSE mineral lineation (Gómez Barreiro 2004).

D$_3$–*M*$_3$

The regional fabrics in the units were overprinted by prograde compressional shear zones and recumbent folds during this phase (Figs 2 and 3). The vergence of the structures and sense of shear were mainly to the NE and east (Martínez Catalán *et al.* 2002; Gómez Barreiro 2004). In the Fornás and Arinteiro units, typical shear zones led to amphibolite-facies re-equilibration with Grt + Pl + Hbl + Zo assemblages. P–T conditions in the highest-grade shear zones were 6–8 kbar and 500–600 °C (Castiñeiras 2003; Gómez Barreiro 2004). Similar conditions were reached in the O Pino unit, where the regional fabric was transposed by a fold-related schistosity and ductile shear zones where assemblages with Grt + Pl + Bt + Qtz were developed in pelitic rocks. Derived P–T conditions there are 7 kbar and 450–550 °C (Gómez Barreiro 2004). This event represents a new cycle of thickening in the units, which is envisaged as a clockwise new loop in their P–T paths (Gómez Barreiro 2004).

Related to this cycle, new extensional structures, such as the Corredoiras and Ponte Carreira detachments (Fig. 1), were developed, representing gravitational readjustments of the tectonic stack (Martínez Catalán *et al.* 1996, 2002; Díaz García *et al.* 1999*b*). The thermal effect of these extensional detachments was restricted to the rock within the shear zone (Díaz García *et al.* 1999*b*; Gómez Barreiro 2004).

D_4–M_4

The D_4 M_4 event was characterized by the development of out-of-sequence thrusts, with a top-to-the-SE sense of shear that reflects the final stage of emplacement of the allochthonous complexes (Martínez Catalán *et al.* 2002; Gómez Barreiro 2004). No thermal effects were detected in the Fornás, Arinteiro and O Pino units, and retrograde metamorphism occurred in the thrust surface, at very low-temperature conditions.

D_5–M_5

Following D_4 contractional event, as well as pervasive shortening and thickening in the underlying autochthon, gravitational disequilibrium led to renewed extensional adjustments.

The Bembibre Pico Sacro detachment is a complex extensional system, which partially reactivated the previous thrusts. Detailed kinematic analysis reveals a shear zone with a general top-to-the-NW sense of shear. The activity of the Bembibre Pico Sacro detachment was contemporaneous with extensive melting in the footwall and the establishment of a high-temperature and low-pressure gradient (Gómez Barreiro *et al.* 2002; Gómez Barreiro 2003), which suggests that the Bembibre Pico Sacro detachment is a structure related to the late gravitational collapse of the orogen (Vanderhaeghe & Teyssier 2001).

D_6–M_6

During this last phase, upright folds with north south axial surfaces were formed, together with strike-slip shear zones and high-angle faults. These features point to a change in the tectonic regime, where transpression represented an important component of shortening (Iglesias & Choukroune 1980; Llana-Fúnez & Marcos 2001; Martínez Catalán *et al.* 2002).

New $^{40}Ar/^{39}Ar$ geochronology

Sample selection

As can be judged from the earlier data, a critical point is to establish the age of the tectonothermal events that followed the maximum pressure, M_1, event in the upper units (Fig. 4). To prevent, as much as possible, thermally induced argon diffusion, we have chosen to focus on the uppermost levels of the HP–HT units, which were expected to have cooled faster than the lower levels during their tectonic exhumation (D_2–M_2; samples IP-05 and IP-08) (House & Hodges 1994; Vance *et al.* 1998). The Fornás and Arinteiro units are a good sample of such levels (Gómez Barreiro 2004). In an attempt to constrain the second cycle of thickening (D_3–M_3), samples were collected from prograde shear zones in the Fornás and Arinteiro (IP-01-03, IP-06 and IP-07) and O Pino (IP-04) units. To investigate the age of the tectonic contacts among different upper units, we have also sampled a prograde shear zone at the bottom of the Monte Castelo unit (IP-11), a gabbroic massif in the NW of the Órdenes Complex occupying a structural position intermediate between the O Pino and Fornás units (Abati 2002), and the Ponte Carreira extensional detachment, at the top of the O Pino unit (IP-09; Fig. 1). A phyllite from the uppermost Betanzos unit (IP-10; Fig.1) has also been sampled, to constrain the age of its tectonic fabric. A brief description of the 10 analysed samples follows.

IP-01 is a garnet amphibolite from a D_3–M_3 prograde shear zone in the Arinteiro unit (Fig. 3). The mylonitic foliation is defined by a syntectonic assemblage composed of Mg-Hbl + Grt + Pl + Zo + Spn + Czo + Qtz + Ilm, and shows a grano-nematoblastic texture. Relics of D_2–M_2 Cpx are preserved as partially resorbed porphyroclasts and lenses, which display a pretectonic relationship with the D_3–M_3 mylonitic foliation. The magnesio-Hbl grain-size fraction (125–62 μm) selected shows textural equilibrium with D_3–M_3 association and defines a strong mylonitic shape fabric of the rocks. Kinematic criteria indicate a top-to-the-NE sense of shear.

IP-03 is a Ca-poor amphibolite, from the same shear zone as IP-01, in the Arinteiro unit (Fig. 3). Mineral composition is Ged + Grt + Qtz + Rt + Ilm +Chl + Prg + St ± Pl ± Bt. The rock shows a nematoblastic texture and mylonitic banding defined by variations on pargasite content. The selected fraction of pargasite (170–250 μm) appears in textural equilibrium with staurolite and gedrite, defining the tectonic shape fabric of the rock. No petrographic evidence of crystal deformation or retrogression has been detected on pargasite.

IP-04 is a garnet amphibolite related to a D_3–M_3 prograde shear zone in the O Pino unit (Fig. 1), containing Mg-Hbl + Pl + Grt + Zo + Rt + Ilm + Qtz. It is a nematoblastic rock with a strong preferred shape orientation. Hornblendes between 75 and 125 μm were selected for experiments. They define the NNE–SSW mineral lineation of the rock.

IP-05 is a low-pressure mylonitic amphibolite (D_2–M_2) from the upper levels of the Fornás unit (Fig. 2), containing Mg-Hbl + Pl + Spn + Ilm ± Qtz. The sample is spatially related to mafic mylonites and ultramylonites with Cum + Act + Pl + Qtz ± Chl ± Rt ± Oam, defining the low-pressure shear zones of the Fornás detachment. Mineral lineation is NNW–SSE, and a top-to-the-NNW sense of shear has been derived from quartz crystallographic preferred orientation and S–C fabrics (Gómez Barreiro 2004). A mixture of Hbl–Pl defines highly strained domains in the rock. Mg–Hbl grain sizes between 75 and 125 μm have been selected from those domains.

IP-06 is a garnet amphibolite from a prograde shear zone (D_3–M_3) in the Fornás unit (Fig. 2). Mylonitic bands are composed of different percentage of syn-D_3 phases: Ts-Hbl + Pl + Ilm + Spn + Grt ± Ep ± Qtz. The rock shows a grano-nematoblastic texture and a strong shape-fabric defined by tschermakitic hornblende (Ts-Hbl). The selected fraction (250–350 μm) of mylonitic hornblendes has no evidence of retrogression or crystal plastic deformation.

IP-07 is a garnet amphibolite from a prograde shear zone (D_3–M_3) in the Fornás unit (Fig. 2), showing grano-nematoblastic texture. A fine mylonitic banding was developed, displaying a homogeneous grain-size distribution. The equilibrium assemblage is Fe-Hbl + Pl + Ilm + Grt ± Qtz. Crystal–plastic deformation was concentrated on plagioclase, with scarce evidence in amphiboles ((010) twins and undulose extinction). Grain sizes between 125 and 250 μm of ferro-hornblende were separated for the experiment.

IP-08 is a high-temperature amphibolite (D_2–M_2) from deep levels of the Fornás unit (Fig. 2). A diffuse gneissic banding is defined by layers with a different mixture of phases Pl + Cpx and Hbl + Cpx ± Pl. The mineral composition is Hbl + Cpx + Pl ± Grt ± Spn ± Cc. The rock shows a granoblastic–grano-nematoblastic texture. The selected amphibole is pargasitic ferro-hornblende (170–250 μm) with evidence of crystal–plastic deformation. The *c*-axes of amphibole and clinopyroxene display strong preferred orientation parallel to the D_2–M_2 NNW–SSE mineral lineation (Van Zuuren 1969; Gómez Barreiro 2004).

IP-09 is a micaschist from the upper levels of the O Pino unit (M_1, St zone). It is deformed by the Ponte Carreira extensional detachment (Fig. 1). Although the mineral association is Ms + Bt + Grt + Qtz, a secondary foliation defined by phengitic Ms, representing neocrystallization of white mica, is the main textural feature in the rock and comprises fractions between 350 and 650 μm, which were selected for single-grain experiments.

IP-10 is a phyllite from the uppermost levels of the upper units of the Órdenes Complex (Betanzos unit) (Fig. 1) collected from the hanging wall to the Ponte Carreira extensional detachment. It has a very fine grain size, and a tectonic foliation parallel to the sedimentary banding (S_0 + S_1) consisting of mica-rich and quartz-rich layers. S_0 + S_1 has been transposed by a continuous crenulation cleavage (S_2). Deformation mechanisms include rigid rotation and recrystallization of micas (sericite, muscovite and chlorite), and pressure solution of quartz. There are no detrital mica relics in the fabric, which displays a homogeneous grain-size population with a greenschist-facies assemblage (Ms + Qtz + Chl). A whole-rock analysis was performed on this sample as no discrete mineral phase could be separated.

IP-11 is a garnet amphibolite from the basal shear zone of the Monte Castelo gabbro (Monte Castelo unit, Abati (2002)) (Fig. 1). It is an example of a prograde D_3–M_3 shear zone from the deep levels of the IP upper units of the Órdenes Complex. The mineral association consists of Ts-Hbl + Pl + Grt + Ilm + Qtz, with the garnet surrounding plagioclase as coronitic aggregates. Textural evidence suggests that crystal plasticity was active during deformation of amphibole (tschermakitic hornblende) and plagioclase. A fraction of amphibole between 170 and 250 μm was selected.

Analytical techniques

Different steps of dating were carried out in the Dep Petrología y Geoquímica, Universidad Complutense, Madrid (sample selection and preparation) and the Department of Isotope Geochemistry, Vrije Universiteit, Amsterdam ($^{40}Ar/^{39}Ar$ laserprobe incremental heating).

Mineral fractions were selected after petrographic analysis. Mineral analyses were performed using a Jeol JXA-8900 M electron microprobe equipped with four spectrometers at the Universidad Complutense. (The analyses are available online at http://www.geolsoc.org.uk/SUP18228. A hard copy can be obtained from the Society Library or from the corresponding author on request.) The operating parameters were 10 s counting time, 15 kV accelerating voltage, 20 nA beam current and a beam diameter between 2 and 5 μm. The ZAF correction procedure was used. To obtain mineral separates, the samples were crushed and sieved as near as possible to the selected fraction. Before final hand-picking, minerals were separated in selected sieve intervals using a Frantz isodynamic separator. Final hand-picking from the mineral separates was carried out under the microscope. Inclusion-bearing grains, crystal fragments and mineral aggregates were discarded. Final concentrates and whole-rock separates were carefully washed with dilute nitric acid and demineralized water. The samples were put into a 3 mm thick tray with 20 holes of 2 mm diameter each. Sixteen holes were loaded with *c.* 0.5 mg of sample. The remaining four positions were loaded with a flux monitor (standard DRA1 sanidine, age 25.26 Ma; Wijbrans *et al.* 1995). A set of trays was irradiated for 12 h with fast neutrons in the CLICIT facility of the TRIGA reactor at the Oregon State University.

The $^{40}Ar/^{39}Ar$ laserprobe incremental heating experiments were carried out with the VULKAAN laserprobe facility at the Vrije Universiteit, Amsterdam (Wijbrans *et al.* 1995). After irradiation, a single fusion experiment was applied to analyse every flux monitor. Irradiation parameter J was calculated for every set of samples: $J = 0.002989 \pm 0.25\%$ for samples IP-01, 06, 08, 10, 11; $J = 0.002929 \pm 0.49\%$ for samples IP-03, 04, 05, 07, 09. The experiments consisted of a total of 6–24 analyses with a stepwise increase of the laser power for every analysis. Incremental heating experiments include 1 min of laser heating with defocused beam, followed by 5 min additional clean-up time per analysis. The system blanks were measured after every three analyses. The isotopic composition of the gas was measured using a double focused noble gas mass spectrometer (MAP 215-50) operating in static mode. Beam intensities were measured on a secondary electron multiplier detector and switchable pre-amplifier resistor settings (10, 100 and 1000 MΩ) by peak jumping at half-mass intervals down from mass 40 to 35.5. The blank correction applied to every unknown was a blank (and its analytical error) that had been integrated to the time of analysis of the unknown after calculating a third-order best fit of all the blank analyses over the day. System mass discrimination was measured by letting clean air Ar (*c.* 4×10^{-15} moles of ^{40}Ar) into the mass spectrometer from an air reservoir (10 l) using a 1 ml gas pipette.

Uncertainties in ages and isotope ratios are reported at 2σ, considering the uncertainties in the determination of the J parameter, the blank determination, the regression of the intensities of individual isotopes, the correction factors for interfering isotopes, and the mass discrimination correction (Wijbrans *et al.* 1995; Koppers *et al.* 2000; Koppers 2002). Data corrected for backgrounds, mass discrimination and radioactive decay are available (see tables in the Supplementary Publication, see above). The use of small mineral fractions leads to a greater analytical error, but gives meaningful absolute ages by preventing analysis of mixed populations (Lips *et al.* 1998). Data processing was carried out using the ArArCALC v2.1.4 software (Koppers 2002).

We use commonly accepted criteria for plateaux as outlined by Koppers *et al.* (2000). The additional criterion not commonly used is that we calculate a mean square weighted deviation (MSWD) the age spectrum, and plateaux will tend to have MSWD <2.0, whereas those plateau segments that do not meet this criterion contain excess scatter, and should be considered as pseudo plateaux or error plateaux. In these cases we believe that the age message may still be meaningful, but the excess scatter precludes a more rigorous interpretation. The full documentation of error handling is beyond the scope of the present paper but has been given by Koppers *et al.* (2000). The calculation strategies for K/Ca profiles and other corrections have been given by Koppers *et al.* (2000) and Koppers (2002). The decay corrected $^{39}Ar_{(K)}/^{37}Ar_{(Ca)}$ ratio was used. A calibration factor was calculated from a hornblende standard with a known homogeneous K/Ca ratio.

Results

The age spectra of samples and K/Ca profiles are shown in Figure 5 (see also Table 1 and the Supplementary Publication, see above), as well as inverse isochron plots for IP-01 and IP-05. Most of the incremental heating experiments resulted in fairly flat spectra with minor to substantial diffusion loss of argon in the lower temperature steps, but in some cases it would appear that several spectra have some minor excess in the first steps rather than diffusion loss. The plateau ages obtained are all interpreted as cooling ages. We have considered a closure temperature of 525 ± 25 °C for hornblende (Harrison 1981), although for amphiboles other than well-crystallized hornblende, lower closure temperatures could be considered. For muscovite, we have used 350 ± 30 °C (Lips 1998; McDougall & Harrison 1999), even when there is some discussion that the closure temperature of muscovite may be somewhat higher: 420 450 °C (Blakenburg *et al.* 1989; Hames *et al.* 1994; Kirschner *et al.* 1996). Others have expressed doubts as to the universal applicability of the closure temperature concept (Villa 1998). The age spectra of samples are shown in Figure 5 (see also Table 1).

The experiments on D_2 M_2 mylonitic fabrics yield well-defined plateaux clustering at 423 382 Ma (samples IP-05 and IP-08). The release spectra are undisturbed and there is no petrographic evidence of D_3 M_3 reworking. Differences in the age could be related to the different structural position of samples during D_2 M_2. K/Ca profiles reflect a homogeneous argon source, with K/Ca values typical of Ca-amphiboles (<0.33) (McDougall & Harrison 1999). The isotope correlation plot for IP-05 reflects the general tendency of all samples in which the range of radiogenic ^{40}Ar ($^{40}Ar_{(r)}$) is between 85 and 100% (see the Supplementary Publication, see above), which causes the points to cluster near the $^{40}Ar/^{39}Ar$ axis, and hence no regression line is defined in the plot. For this reason further isotope correlation plots are not included.

The amphibolites and metasediments with D_3 M_3 mylonitic fabrics yield well-defined plateau ages between 410 and 371 Ma in samples IP-01, 03, 04, 06, 07, and 09. Sample IP-11 has an irregular saddle-shaped spectrum. The initial argon release yields old apparent ages, which decrease with progressive ^{39}Ar release, defining an imprecise plateau of 393 ± 35 Ma, which probably represents cooling after D_3 thrusting, and there is a return to older ages at the end of the spectrum. Special

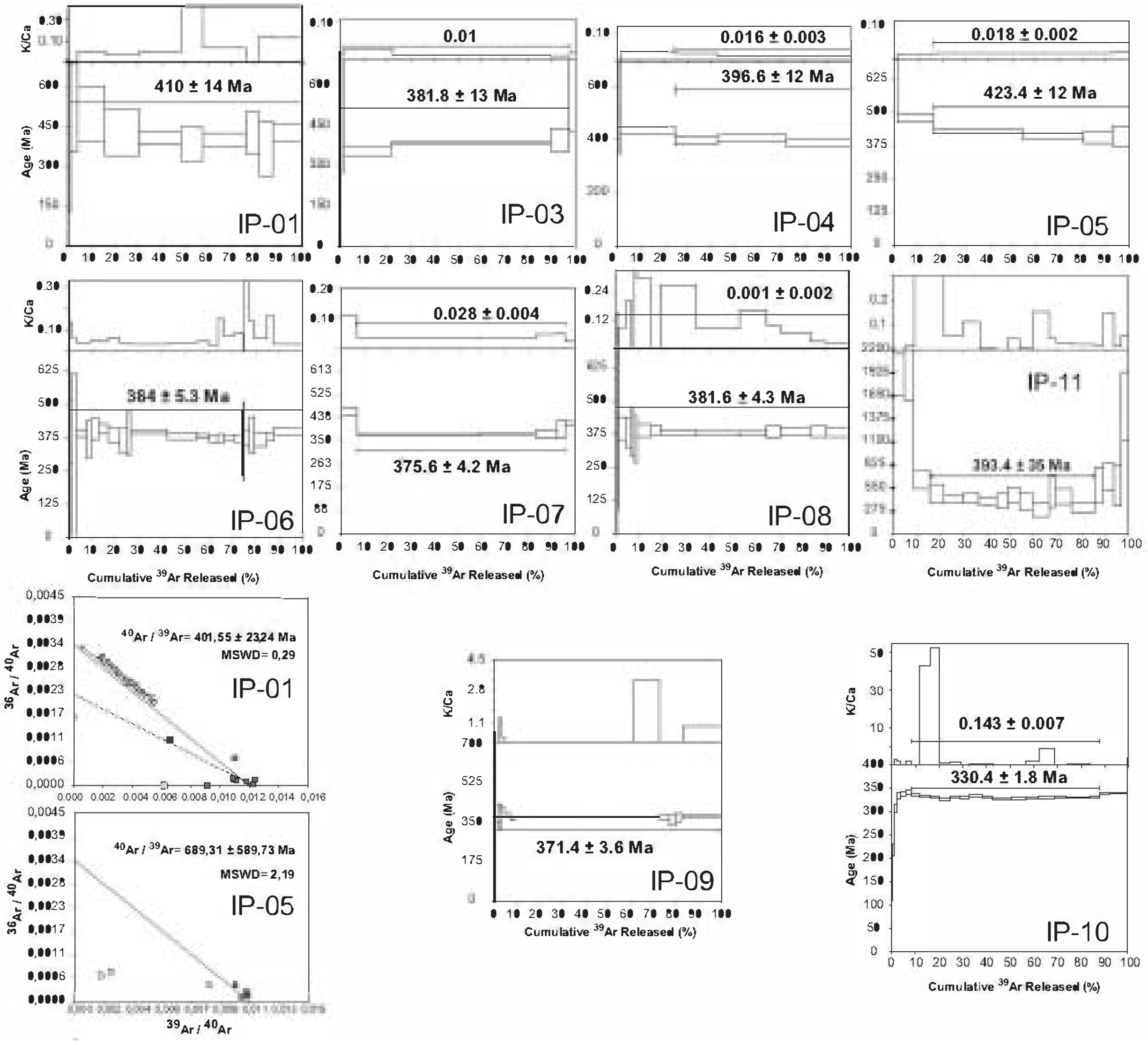

Fig. 5. Results of ^{40}Ar/^{39}Ar laserprobe dating. All age spectra and K/Ca profiles are represented with a 2σ analytical error, and are summarized in Table 1 and the Supplementary Publication (see p. 67). Isotope correlation plots for IP-01 and IP-05 are shown; white boxes represent rejected steps. (Structural position of samples is shown in Figs 1–3.)

diffusion mechanisms, such as anion diffusion, have been invoked to explain U-shaped release patterns (Harrison & McDougall 1981), but recent data suggest that fluid inclusions, released at low temperature, and solid inclusions, released at high temperature during the experiment could be the most likely explanation (Esser *et al.* 1997; Boven *et al.* 2001; Kelley 2002). In general, K/Ca profiles reflect a homogeneous source of argon, and no evidence of phase mixture exists (K/Ca <0.33

Table 1. *Summary of ^{40}Ar/^{39}Ar laserprobe dating results*

Sample	Rock type	Structural unit	Type of analysis	Apparent total gas age (Ma, 2σ)	Plateau age (Ma, 2σ)	% ^{39}Ar (in plateau)
IP-01	Grt-amphibolite	Fornás–Arinteiro (HP–HT)	Hbl	438.9 ± 25	410 ± 14	98.98
IP-03	Ca-poor amphibolite	Fornás–Arinteiro (HP–HT)	Hbl	390.4 ± 10	381.8 ± 13	97.10
IP-04	Grt-amphibolite	O Pino (IP)	Hbl	416.4 ± 8.4	396.6 ± 12	75.37
IP-05	Low-*P* amphibolite	Fornás–Arinteiro (HP–HT)	Hbl	440.3 ± 8.7	423.4 ± 12	83.5
IP-06	Grt-amphibolite	Fornás–Arinteiro (HP–HT)	Hbl	391.6 ± 8.7	384 ± 5.3	99.77
IP-07	Grt-amphibolite	Fornás–Arinteiro (HP–HT)	Hbl	387.3 ± 4.5	375.6 ± 4.2	88.76
IP-08	High-*T* amphibolite	Fornás–Arinteiro (HP–HT)	Hbl	388.6 ± 5.7	381.6 ± 4.3	98.64
IP-09	Micaschist	O Pino (IP)	Ms	371.3 ± 3.8	371.4 ± 3.6	99.91
IP-10	Phyllite	Betanzos (IP)	WR	330.2 ± 1.6	330.4 ± 1.8	79.45
IP-11	Grt-amphibolite	Monte Castelo (IP)	Hbl	676.2 ± 30.7	393.4 ± 35	70.44

in Ca-amphiboles), except for low-energy steps where excess argon has been detected (e.g. IP-11). Other variations along the K/Ca profile (IP-06) are irrelevant from a statistical point of view at the time of age calculations (Fig. 5).

Muscovite analysis (IP-09) shows a flat age spectrum, which is considered as evidence of an undisturbed system. There is no evidence of argon loss in the analysis or polyphase mixtures (McDougall & Harrison 1999).

A considerably younger age was obtained from the phyllite of the Betanzos unit. The whole-rock spectrum of sample IP-10 yields a concordant plateau of roughly 330 ± 2 Ma that clearly points to the Variscan cycle in the strict sense. The K/Ca profile is consistent with petrographic, chemical and spectrum data (see the Supplementary Publication, see p. 000). The small grain-size of the sample precluded mineral separation. In our case, isotope composition of extracted gas reflects no relevant phase mixture, and hence it is considered a meaningful age (McDougall & Harrison 1999).

Discussion

Age of the tectonic exhumation D_2–M_2

The Fornás detachment is a major extensional contact between the HP–HT (Fornás and Arinteiro) and IP (O Pino) units. Sample IP-05 was collected in an extensional shear zone developed during the last stage of exhumation (*c.* 550 °C), from the upper levels of the Fornás HP–HT unit, whereas sample IP-08 represents the higher temperature decompression stage (*c.* 750 °C) and is located deeper in the same unit. Structural position may be an important factor to explain age differences between the two samples. No statistical dependence has been found between the age of the amphiboles and composition (e.g. Mg-number, K/Ca, K_2O) as has been suggested by some workers (e.g. Dahl 1996; Reddy *et al.* 1997; McDougall & Harrison 1999), and specific experiments should be designed to check this possibility in the future.

Data from other low-dipping extensional structures indicate that in the footwall to a detachment, rocks near the detachment surface underwent higher cooling rates than rocks away from it, resulting in $^{40}Ar/^{39}Ar$ apparent ages progressively older as the detachment is approached (e.g. House & Hodges 1994; Vance *et al.* 1998). Differences in age may reflect that the upper level of the footwall (sample IP-05, 423 ± 12 Ma) was cooling faster than the lower levels (sample IP-08, 382 ± 4 Ma), as no analytical evidence of excess argon exists. This suggests that the activity of the Fornás detachment started at least before 423 ± 12 Ma. This conclusion supports previous geochronological results in other HP–HT units of the Órdenes and Cabo Ortegal complexes (Fig. 6), which yield similar ages (425–410 Ma) for the tectonic exhumation stage (Dallmeyer *et al.* 1997; Fernández-Suárez *et al.* 2004). This interval is considered a minimum age for the beginning of tectonic exhumation, and also the HP–HT stage (D_1–M_1) that preceded it. Figure 6 shows a summary of data for this interval (D_2–M_2) consistent with the proposed interpretation. In some cases, lack of detailed tectonometamorphic characterization of geochronological samples precludes a more exact location of results in the figure (e.g. Peucat *et al.* 1990; Ordóñez Casado *et al.* 2001). On the other hand, a detailed experiment conducted on HP–HT units (Fernández-Suárez *et al.* 2004) revealed a complex evolution of zircons in these rocks. These facts highlight the importance of improving the textural control of mineral phases used in U–Pb geochronology before deriving a definite interpretation of age data (e.g. Rubatto 2002).

Age of the D_3–M_3 episode

The ages obtained from prograde compressional shear zones (IP-01–IP-04, 06, 07) and subsequent extensional detachments (IP-09) are very consistent across the IP and HP–HT units in the Complex (Fig. 6). Plateau ages between 410 and 371 Ma confirm the existence of a regional tectonothermal event related to the Variscan convergence, in which an Early Variscan stage has been identified, and correlate well with the Acadian event proposed by some workers in the Cabo Ortegal Complex (Ordóñez Casado *et al.* 2001). In the Órdenes Complex during this stage prograde, amphibolite-facies shear zones and recumbent folds were developed. In the Cabo Ortegal Complex, *P–T*

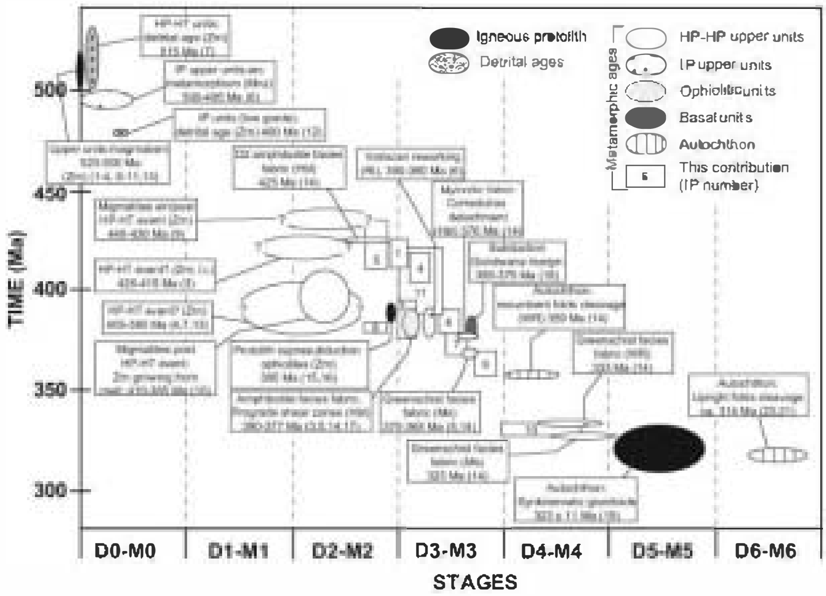

Fig. 6. Summary of representative ages from the NW Iberian Massif, from upper allochthonous units to the autochthon sequences. (See the text for a discussion.) Zrn, zircon; Mnz, monazite; Ms, muscovite; Hbl, hornblende; WR, whole-rock analysis; l.i., lower intercept. References: (1) Van Calsteren *et al.* (1979); (2) Kuijper (1979, 1980); (3) Peucat *et al.* (1990); (4) Schäfer *et al.* (1993); (5) Dallmeyer *et al.* (1991); (6) Abati *et al.* (1999); (7) Ordóñez Casado *et al.* (2001); (8) Santos *et al.* (2002); (9) Fernández-Suárez *et al.* (2002); (10) Fernández-Suárez *et al.* (2004); (11) Dallmeyer & Tucker (1993); (12) Fernández-Suárez *et al.* (2003); (13) Santos Zalduegui *et al.* (1996); (14) Dallmeyer *et al.* (1997); (15) Pin *et al.* (2002); (16) Díaz García *et al.* (1999*a*); (17) Dallmeyer & Gil Ibarguchi (1990); (18) Rodríguez *et al.* (2003); (19) Bellido *et al.* (1992); (20) Capdevila & Vialette (1970); (21) Ries (1979).

conditions could be higher, a fact that may be related to different positions in the orogenic wedge (Gómez Barreiro 2004), or included into a single orogenic cycle model of Variscan age (Ordóñez Casado *et al.* 2001), rejecting any evidence that would locate the D_2–M_2 stage in pre-Variscan times as we have found.

In our samples, tectonometamorphic evidence indicates that these ages do not represent the HP–HT episode (M_1), but clearly postdate it, giving rise to the polyorogenic character of the upper units. Also, the activity of extensional detachments during this stage has been confirmed. An age of 371 ± 4 Ma has been obtained for the Ponte Carreira extensional detachment reflecting the cooling of mylonitic fabrics below 350°C. A similar structure in the Órdenes Complex, the Corredoiras detachment (see Fig. 1), has yielded an argon age of *c.* 375 Ma (Dallmeyer *et al.* 1997).

Age of the D_4 M_4 episode

The 330 Ma obtained from a low-grade phyllite of the Betanzos unit is comparable with the 333 Ma obtained in similar rocks by Dallmeyer *et al.* (1997), and related by those workers to late upright folding and the thermal imprint of syntectonic granitoids. However, it is somewhat old for the age of steep folds in the autochthon (314 ± 6 Ma; Capdevila & Vialette 1970; Ries 1979) and also for the synkinematic granitoids a few kilometres to the east (323 ± 11 Ma; Bellido *et al.* 1992). On the other hand, the age obtained fits well into the interval where the out-of-sequence thrusts were active and upright folds developed in their hanging-wall units by duplexing (Martínez Catalán *et al.* 2002). We consider that the *c.* 330 Ma ages could represent the D_4–M_4 event, either by cooling related to exhumation by thrusting, or as resetting ages associated with foliation development during contemporaneous folding.

Tectonothermal evolution of the upper units

Tectonometamorphic and age data seem to fit into a three-stage model (Fig. 7) as follows.

(1) In a first cycle of burial by subduction, crustal thickening and exhumation (D_1–M_1 and D_2–M_2), maximum pressures were reached in the IP and HP–HT units. Extensional fabrics related to the tectonic exhumation (D_2–M_2), yielding ages of 425–410 Ma. These data correlate with other metamorphic ages in similar allochthonous terranes in Europe (Oliver *et al.* 1993; von Quadt & Gebauer 1993; Ballèvre *et al.* 1994; Timmermann *et al.* 2004) and in the North America counterpart in Newfoundland (Dunning *et al.* 1990; Cawood *et al.* 1995). Taking into account that in high-pressure units, exhumation closely follows burial, and that the 425–410 Ma U–Pb and $^{40}Ar/^{39}Ar$ ages reflect exhumation of the HP–HT units, a Silurian–Early Devonian age is reasonable for this cycle. In the Early Silurian, Avalonia accreted to Laurentia–Baltica forming Laurussia (Lefort 1989; Murphy *et al.* 2004), and this probably was the large continental mass to which the Iberian upper allochthon accreted. This first cycle is considered pre-Variscan because it is related to the closure of either the Iapetus or the Tornquist ocean, and not to that of the Rheic ocean.

(2) A second cycle of burial and exhumation (D_3–M_3) shows a diachronous evolution across the allochthonous units in NW

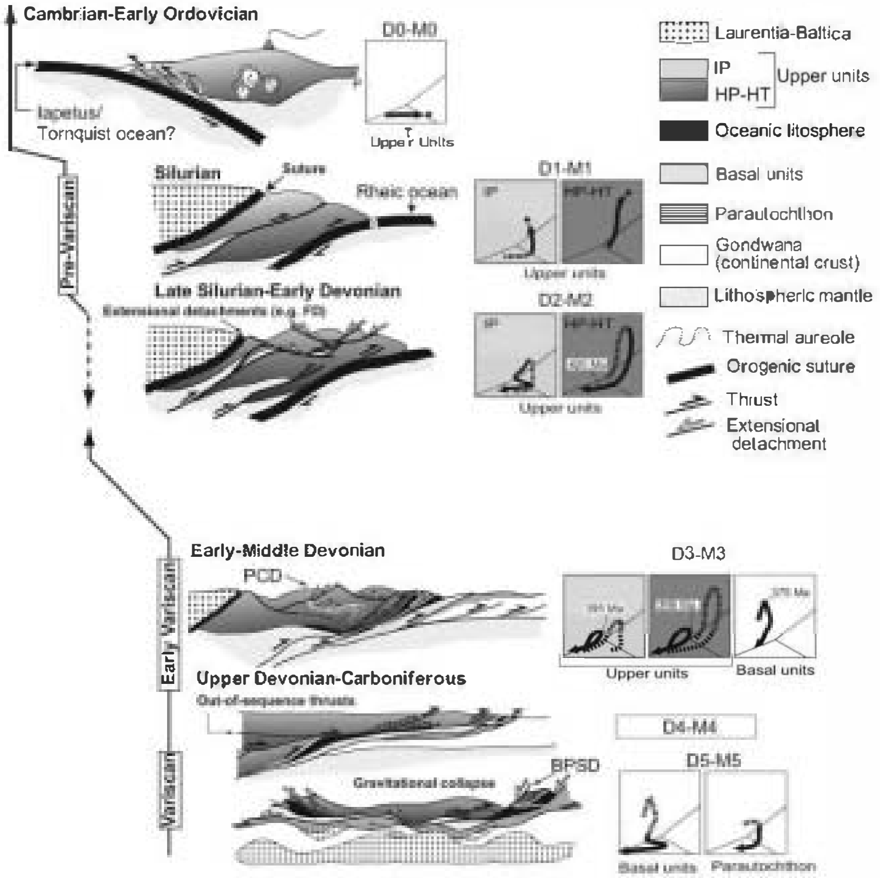

Fig. 7. Tectonothermal evolution of the allochthonous units of the Órdenes Complex according to the proposed model of three cycles of convergence and exhumation. (1) Pre-Variscan convergence: D_0 M_0, formation of a volcanic arc at the NW active margin of Gondwana. D_1 M_1, accretion of the arc to the margin of the Rheic ocean opposite to Gondwana. D_2 M_2, tectonic exhumation of the units by extensional detachments. (2) Early Variscan convergence: D_3 M_3, continued development of the orogenic wedge by understacking of the ophiolitic and basal units, resulting in a new stage of burial and exhumation. (3) Variscan convergence: D_4 M_4 and D_5 M_5 represent thes final stages of emplacement of the allochthonous complexes, including the gravitational collapse of the orogen. Abbreviations are as in Figure 1.

Iberia (Dallmeyer *et al.* 1997). In the case of the upper allochthonous units this cycle started after 410 Ma, probably beginning in the Early Devonian and spanning the Mid-Devonian. The related convergence led to the closure of the Rheic ocean at *c.* 380 Ma (Dallmeyer & Gil Ibarguchi 1990; Dallmeyer *et al.* 1991) and subduction of the Gondwana margin at *c.* 375 Ma (Van Calsteren *et al.* 1979; Rodríguez *et al.* 2003). This cycle is considered early Variscan because it is related to Laurussia Gondwana plate convergence and the closure of the Rheic ocean, but its age (400 370 Ma in the upper allochthon, up to 360 Ma in the ophiolites and basal allochthon; Dallmeyer *et al.* 1997; Rodríguez *et al.* 2003) is older than those normally considered Variscan in a strict sense, that is, essentially Carboniferous.

(3) A renewed cycle of crustal thickening (D_4 M_4) placed the allochthonous complexes in its present position, above the sequences of the passive margin of Gondwana, and was followed by final gravitational collapse (D_5 M_5) and residual contractional tectonics producing upright folding and transcurrent shearing (D_6 M_6). This third cycle is clearly Variscan and extends from the Late Devonian throughout the Carboniferous. It is the only cycle responsible for the deformation and metamorphism in the parautochthon and autochthon, where it also evolved sequentially and diachronously, prograding toward the external zones with time (Dallmeyer *et al.* 1997).

Conclusions

The age of early tectonothermal events in a polyphase allochthonous terrane of the Iberian Variscan belt has been investigated by means of $^{40}Ar/^{39}Ar$ laserprobe incremental heating experiments, to constrain its accretionary history. Dating of fabrics linked to structures whose meaning and relative ages are well constrained reveals a pre-Variscan accretionary stage, in addition to well-known Early Variscan and Variscan events.

Two groups of ages have been defined taking into account the tectonothermal sequence: (1) Silurian Early Devonian, obtained from the mylonites of the Fornás extensional detachment, here considered as the minimum age for the start of tectonic exhumation of HP HT units and an upper limit for the HP HT event itself; (2) Early to Mid-Devonian, derived from structures related to the Early Variscan convergence in the area. A single, younger Carboniferous age possibly reflects the final stages of emplacement of the allochthonous complexes.

Our data offer a basis to interpret a part of the Iberian allochthonous complexes as polyorogenic, and evolved through three complete cycles of burial, thickening and exhumation, all of them related to the Laurussia Gondwana convergence during the Palaeozoic. The older, Early Palaeozoic cycle has been identified only in the upper allochthonous units, remnants of a peri-Gondwanan terrane with Cambro-Ordovician arc magmatism that overlie ophiolites derived from the Rheic ocean. Early Palaeozoic convergence was responsible for the accretion of this terrane to either Laurentia or Baltica prior to the closure of the Rheic ocean. Depending on the unconstrained position of Iberia at that time, that convergence may be related to the closure of either the Iapetus or the Tornquist ocean. The second cycle is related to the closure of the Rheic ocean, and its effects can be seen in all allochthonous units. The third is a product of the Laurussia Gondwana collision, and involved both the allochthon and the autochthon.

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